A Coupled Atmosphere–Ocean Model in the Tropics with Different Thermocline Profiles

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ABSTRACT

A coupled atmosphere–ocean model is used to investigate some important effects of a steep sloping thermocline in the central Pacific. It is found that the coupled modes are very sensitive to the steepness of the thermocline in the central Pacific Ocean. The wave reflection and modal decomposition processes play an important role in the initial development of the unstable mode and dramatically affect the fates of the oscillation. A sensitivity test is conducted to test the role of western boundary reflection in this particular model. The insensitivity of the western boundary reflection seems to agree with the results of previous research that showed that coupled unstable modes do not necessarily depend on Rossby wave reflections.

1. Introduction

The intriguing hypothesis of Bjerknes that an El Niño–Southern Oscillation (ENSO) event results from the positive feedback between the ocean and the atmosphere has stimulated studies of air–sea interactions in the tropics and led to successful ENSO predictions by using a simple coupled model (Cane et al. 1986; Zebiak and Cane 1987). However, long-term ENSO prediction can be improved by using appropriate model initialization and background specification. The important dependence of the ENSO-like oscillation on a model background was demonstrated by Neelin (1990) and Wakata and Sarachik (1991) who showed that a change of air–sea coupling intensity could dramatically change the oscillation patterns. The intensity of air–sea interaction strongly depends on the local background condition, such as climatological wind speeds and SST. In this work, we attempt to study some effects of one nonuniform feature in the tropical ocean, the zonally sloping thermocline, on the evolution of ENSO-like coupled modes.

In response to the atmospheric forcing, both the climatological SST and the thermocline depth increase westward in the equatorial Pacific. The Wentzel–Kramers–Brillouin–Jeffreys (WKBJ) method has been used to estimate the changes of wave properties in a gently varying thermocline (e.g., Hughes 1981; Yang and Yu 1992). This condition is invalid in the central Pacific where the thermocline depth changes rapidly. Gill and King (1985) used a two-layer reduced-gravity model to study how such a thermocline ramp affects wave propagation. For a low-frequency Kelvin wave, the sloping thermocline acts like a step. Therefore the dynamic matching condition requires that up to 25% of the incident Kelvin wave’s energy be reflected westward as long Rossby waves. Analytic study conducted by Bussalachi and Cane (1988) found similar results. Another main result of Gill and King (1985) is that the eastern Pacific Ocean is not necessarily dominated by variability associated with the first baroclinic mode even if the remote signals in the western Pacific Ocean are completely dominated by the first mode. The vertical modal energy exchanges can be very efficient for low-frequency Kelvin waves when they propagate through a steep thermocline in the central Pacific.

If the ocean were free from interaction with the atmosphere, the results of Gill and King (1985) might suggest that a thermocline front is not an effective barrier for zonal energy propagation. However, the subsequent air–sea interaction processes resulting from Rossby wave reflection and Kelvin wave decomposition, modal decomposition, defined by Gill and King (1985), refers to the energy exchanges between different vertical modes, could be vitally important in determining the evolution of an unstable coupled mode. As we will discuss later, the thermocline front can reduce the intensity of tropical air–sea interactions by several processes. First, Rossby wave reflection reduces the SST anomaly in the eastern Pacific Ocean. Second, the wind fields induced by different Kelvin modes and reflected Rossby waves damp each other.

The model formulation is given in section 2. The model results will be presented and discussed in section 3. We will also test the role of western boundary reflection in this model.

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2. Model formulation

a. The ocean model

A 2/2-layer reduced-gravity ocean model is used. The depth of the upper-layer interface is shoaling eastward to represent the depth of 20°C isotherm in the equatorial Pacific Ocean, while the lower-layer interface is flat at the 10°C isotherm. Advection due to the mean currents is ignored in this work. Since our purpose is not to simulate the ENSO events with all realistic features, we exclude the mean-flow effects in our model for the sake of simplicity. We must point out that the exclusion of the mean currents are not rigorously justified.

The equations that govern the two layers are:

\[
\begin{align*}
\partial_t u_1 + \beta y v_1 &= -p_{1x} - \gamma u_1 + r^*/(H \rho_1), \\
\partial_t v_1 + \beta y u_1 &= -p_{1y} - \gamma v_1 + r^*/(H \rho_1), \\
\partial_t h_1 + (H u_1)x + (H v_1)y &= 0,
\end{align*}
\]

where \((u_1, v_1)\) and \((u_2, v_2)\) are velocity components in the upper and lower layers, \(h_1\) and \(h_2\) are anomalous-layer thicknesses, \(H = H_1 + H_m\) is the mean upper-layer thickness (including the mixed layer), \(H_2\) is the mean layer thickness of the lower active layer, and \(-\gamma(u_1, v_1)\) and \(-\gamma(u_2, v_2)\) represent the Rayleigh friction. The wind stress is estimated by the aerodynamic bulk formula and the same drag coefficient as Zebiak (1984) is used.

The pressure gradients in the two upper layers are given by

\[
\nabla p_1 = \left(\frac{\rho_2 - \rho_1}{\rho_3}\right) g \nabla h_1 + \left(\frac{\rho_3 - \rho_2}{\rho_3}\right) g \nabla h_2,
\]

\[
\nabla p_2 = \left(\frac{\rho_2 - \rho_1}{\rho_3}\right) g \nabla h_1 + \left(\frac{\rho_3 - \rho_2}{\rho_3}\right) g \nabla h_2,
\]

where \(\rho_1, \rho_2,\) and \(\rho_3\) are the water densities in the upper, lower, and deep layers, respectively, and \(g\) is the gravitational acceleration rate. The density differences between layers are chosen as \((\rho_2 - \rho_1) = \alpha \rho_0 \Delta T_1\) and \((\rho_3 - \rho_2) = \alpha \rho_0 \Delta T_2\), where \(\alpha = 0.00025^\circ C^{-1}\) is the thermal expansion coefficient, \(\Delta T_1 = \Delta T_2 = 10^\circ C\) are the temperature differences between two layers, and \(\rho_0 = 1020\) kg m\(^{-3}\) is water density. Following Zebiak (1984), a constant-depth mixed layer is added to improve the model estimate of surface dynamics and thermodynamics. Figure 1 schematically shows the model formulation. In this model, we choose \(\gamma = 1/250\) days, \(H_m = 50\) meters, and \(H_0 = H_m + H_1 + H_2 = 350\) meters. Profiles of \(H(x) = H_1(x) + H_m\) for different simulations will be discussed separately.

The mixed-layer formulation of Zebiak (1984) is adopted in this work. For detailed derivation of mixed-layer velocity \((u_m, v_m)\), and entrainment rate \(w\), one may refer to Zebiak (1984). The SST evolution is determined by horizontal advectons, vertical entrainment, and surface fluxes. The SST equation is very similar to that of Zebiak (1984) and Zebiak and Cane (1987), that is,

\[
T_t = -u_m V(T_0 + T) - w_0 VT
\]

where

\[
M(W_0 + w) - M(W_0) \frac{T_0 - T_{sub}}{H_m} - M(W_0 + w) \frac{T - T'_e}{H_m} - \alpha T
\]

is the Heaviside step function, \(T_{sub}\) is the mean subsurface water temperature, \(T'_e\) is the anomalous subsurface water temperature, and \(-\alpha T\) is Newtonian cooling, which represents a linearization of surface heat flux. Yang (1991) showed that the advection due to mean flow, \(u_0 VT\), is considerably smaller than the advection due to the anomalous currents, \(u_m VT_0\), which is consistent with the scale-analysis result conducted by Neelin (1991). Hence, we neglect \(u_0 VT\) in our model for the sake of simplicity although such simplification is not rigorously justified.

In (3) \(V T_0\) is estimated by using the annual mean SST of Shea et al. (1990) and \(\alpha = 1/(200\) days) is used as the coefficient of Newtonian damping. A uniform value of the mean equatorial upwelling rate \(W_0 = 1.15 \times 10^{-5} \) m s\(^{-1}\) or 1 m day\(^{-1}\), as estimated by Wyrtki (1981), is used in this model.

Zebiak and Cane (1987) parameterized the anomalous temperature as \(T'_e = \theta(h) \tanh(\lambda(H + 1.5|h|))\)
\[\tanh(\lambda h)\], where \(\theta = 28\ \text{K} \text{ and } \lambda^{-1} = 80\ \text{m} \text{ for } h > 0, \text{ and } \theta = -40\ \text{K} \text{ and } \lambda^{-1} = 80 \text{ for } h < 0. \text{ In this model, we use a different formula derived from observational data (Levitus 1982) by Seager et al. (1988), who used a cubic spline interpolation to derive a relationship between the subsurface temperature at 50 meters below the sea surface and the depth of the 20°C isotherm. This formula shows that the subsurface temperature increases rapidly as the main thermocline deepens, and is much more sensitive to thermocline depth changes in the eastern Pacific where the thermocline is shallower.}

In Zebiak (1984) and Zebiak and Cane (1987), the anomalous entrainment by the mean upwelling was parameterized as
\[-[M(W_0 + w) - M(W_0)] \frac{T_0 - T_{\text{sub}}}{H_m}\]
\[= -[M(W_0 + w) - M(W_0)] T_{0z},\]
where \(T_{0z}\) is a specified vertical gradient of the mean temperature. In our model, a different parameterization is used, that is,
\[-[M(W_0 + w) - M(W_0)] \frac{T_0 - T_{\text{sub}}}{H_m}\]
\[= -[M(W_0 + w) - M(W_0)] \frac{T_0 - T_e(H)}{H_m}, \quad (4)\]
where \(T_e(H)\) is the water temperature at the base of the mixed layer and \(H\) is the mean thermocline depth.

\[b. \text{ The atmospheric model}\]

A steady-state atmosphere model (Gill 1980, 1982) is used to calculate anomalous wind speeds:
\[-\beta y V = -P_x - \gamma U, \quad (5a)\]
\[\beta y U = -P_y - \gamma V, \quad (5b)\]
\[C^2(U_x + V_y) = -Q - \gamma P, \quad (5c)\]

where \(Q\) is the forcing term relating to latent heating, the last terms of (5a)-(5c) represent the Rayleigh friction and Newtonian damping, and \(C\) is the internal gravity wave speed. Discussion of using such a simple model can be found in Gill (1982) and Zebiak (1982).
In this study, we choose $C = 60 \text{ m s}^{-1}$ and $\gamma = 1 (5 \text{ days})^{-1}$.

The anomalous diabatic heating term $Q$ in (5c) is estimated by using the Clausius–Clapeyron equation (Battisti 1988).

3. The model results and discussion

The model is discretized into an Arakawa C grid. The model width is taken to be 160° in latitude with a grid size of 1.5°. The meridional boundaries, located at 40°S and 40°N, are open, and the method of Camerlengo and O'Brien (1980) is imposed to compute model variables along the boundaries. No normal-flow and no-slip boundary conditions are used along the solid boundaries (Yu et al. 1991).

a. Results

Three numerical experiments have been performed and results of each experiment will be discussed. The same model parameterization and initial condition were used in all experiments. In the first experiment a more realistic thermocline profile is used (solid line in Fig. 2). In this experiment the thermocline depth in the western and central Pacific Ocean from 130°E to 170°W is 175 m below the sea surface. It decreases linearly from 175 m at 170°W to 75 m near the eastern boundary at 110°W, and remains constant east of 110°W. In the second experiment, a steeper thermocline is confined to a small area from 130°W to 110°W. The third experiment is conducted by using a flatter thermocline profile, where it increases from 200 m at 170°E to 75 m at 110°W.

The initial condition was created by forcing the ocean model with a westerly wind patch in the western Pacific Ocean for 20 days. In response to such atmospheric forcing, downwelling Kelvin waves and upwelling long Rossby waves are generated. In all the experiments the ocean and the atmosphere communicate with each other once a day, that is, the anom-
alous wind stress is updated every model day and remains the same between the model days.

For the first experiment, Fig. 3a shows the anomalous SST evolution at the equator off the eastern boundary over a period of ten years. A regular interannual oscillation with a period of about 2 ~ 3 years is self-sustained. The maximum SST anomaly is about 3°C in the eastern equatorial Pacific Ocean. The anomalous depth of the main thermocline is depicted in Fig. 3b. The anomalous thermocline depth, having an amplitude of about 35 m, also oscillates with the same period. The second-layer thickness anomaly is shown in Fig. 3c.

The eastward propagation of this oscillatory mode is evident in a longitude vs time plot of anomalous SST (Fig. 4a). The initial anomalous SST propagates eastward with increasing amplitude due to positive feedback from anomalous winds. The SST warms up rapidly near the eastern boundary between 110°W and 70°W. At about the 200th day the warm event begins to decay while a negative anomalous SST starts to develop west of the warm event. The cooler water slowly intrudes eastward and cools the previously existing warm SST in the eastern ocean. At the 500th day the warm SST anomaly completely vanishes and the eastern Pacific Ocean starts to develop a cold phase of the SST. It is interesting to note that a positive SST anomaly starts to develop and slowly migrate eastward when the cold event begins to amplify in the eastern Pacific. This weak warm signal amplifies quickly in the eastern ocean at 110°W. At about the 1000th day this warm event completely dominates the entire eastern equatorial Pacific Ocean. The model ocean continues to oscillate in a similar manner in the remaining time. Figure 4a shows that the model experiences four warm events and three cold events in a period of ten years.

The evolution of the anomalous thermocline depth in the eastern equatorial Pacific is very similar to the anomalous SST (Fig. 4b). The origins of the warm and cold events in the western and central equatorial
Pacific Ocean, however, are more evident in the contours of the anomalous thermocline depth. For example, during the early stage of the first warm event, an upwelling signal associated with a cold event is clearly detected in the western Pacific between 130°E and 170°E. The upward thermocline depth deviation of this potential cold event has an amplitude of over 10 m. When this upwelling signal moves eastward to the central equatorial Pacific Ocean, a downwelling wave has already formed near the western boundary. An upwelling Kelvin wave is also detected during the opposite phase of the oscillation between the 900th and 1400th day. The model equatorial Pacific Ocean always experiences two conditions along the equator. When an El Niño develops near the eastern boundary, upwelling processes are dominated in the western and central Pacific and slowly intrude eastward to terminate the El Niño. The SST anomalies of this oscillatory mode are mainly localized in the eastern Pacific Ocean. The physical reason for this SST localization is due to the inhomogeneous mean background in the zonal direction. In the western Pacific Ocean, only small SST anomalies will be induced by finite amplitude waves because of the small temperature gradients in the vertical and horizontal directions; hence, the oceanic thermodynamics are less influenced by those disturbances in the western Pacific.

Now consider a situation in which the thermocline slope in the central equatorial Pacific Ocean is steeper. Figure 5a shows the anomalous SST evolution just off the eastern boundary at the equator. The initial disturbance does reach the eastern boundary and causes a local warming there. The SST oscillation, however, is damped, and the model returns to the equilibrium state of rest after six years. Although the steep thermocline does not terminate the initial development of the warming in the eastern Pacific Ocean, the amplitude of the anomalous condition is reduced to about 2.3°C compared with the initial peak of 3°C in the first experiment. It is likely that the reduction of the
oscillation amplitude in the first cycle strongly influences the further evolution. The maximum SST anomaly of the following cold event is further reduced. Figure 5a shows that the SST in the eastern ocean is about 1.5°C below the mean condition, which is smaller than 3.85°C in the first experiment (Fig. 4a). Under this condition the model experiences two weak warmings and two very weak coolings before it returns to the state of rest. The deviation of the thermocline depth and the second-layer thickness anomalies also damp in the same time scale. This experiment suggests that the mean background specified in this experiment is unfavorable to sustain an oscillatory mode.

Next we will discuss the model results when the thermocline profile is flatter than the “normal condition.” The flatter thermocline slope in this experiment reduced the reflection of long Rossby waves and the degree of the vertical modal decompositions. Figure 6a shows anomalous SST versus time. The amplitude of the SST anomaly gradually increases from 3.15°C in the peak of the first warm event to 4.5°C in the peak of the second warming. The model develops the numerical instability at the beginning of ninth model year. The numerical instability is due to the surfacing of the interface between the two active layers in the shallow area in the eastern Pacific Ocean. This can be seen in Fig. 6b, which shows the anomalous depth of the interface. During the peak of the cold event, the upward deviation of the main thermocline is about 75 meters, which is about the mean thermocline depth specified in the eastern ocean. Strictly speaking, the formulation of the mixed layer, during a peak of the cold phase of ENSO, does not hold because the thermocline has almost surfaced, and the mixed layer should be very shallow or even vanish locally. However, we still adopt this model formulation in this special case for the purpose of allowing direct comparison with the results of the “normal case” (experiment one). The second-layer thickness anomaly evolution is plotted in Fig. 6c. The period of the oscillation is slightly shorter than that of the first experiment.

In these three experiments, the condition of the first experiment is regarded as a “normal condition” of the Pacific Ocean where the interannual oscillation is self-sustained. If the Pacific Ocean possessed a condition similar to that in the second experiment, continuous oscillations could be more difficult to sustain. If the thermocline slope barrier were weaker, as in the third experiment, stronger oscillation could be sustained unless some limiting factors, such as nonlinearity, were included. The ocean condition in the Pacific is varying. Some of these variations are in part associated with ENSO events, and some are attributable to the extratropical influences. If the oceanic condition just before the onset of an El Niño event had drifted from a normal condition to a condition similar to experiment 2, it would be unfavorable for the further development of ENSO. Under this condition, one might expect a smaller El Niño or an aborted El Niño. If the oceanic condition deviated from its seasonal cycle toward a favorable condition, as in the third experiment, the oscillation within this cycle might have a greater amplitude.
FIG. 6. The evolution of the model variables in the eastern boundary at the equator: (a) the SST anomaly; (b) the upper-layer thickness anomaly; (c) the second-layer thickness anomaly (run 3).

b. The role of thermocline reflection and modal decomposition

Does the wave reflection or modal decomposition play any role in determining the evolution of these oscillatory modes? Some previous works, such as Gill and King (1985) and Long and Chang (1990), already showed that thermocline reflection or modal decomposition is very effective to uncoupled modes. To the coupled atmosphere–ocean system, the feedbacks between reflected and decomposed modes could provide additional dampings. Such damping mechanisms can be schematically explained in Fig. 7, which shows the anomalous winds induced by one and two positive SST patches. In Fig. 7a, a positive SST anomaly causes a local atmospheric convection with easterly winds on the eastern side and westerly winds on the western side of the SST. The anomalous winds force the ocean and provide a positive feedback to allow coupled instability to develop. If the SST is broken into two separate SST patches, as shown in Fig. 7b, each SST anomaly tends to cause its own convergent winds, and the westerly induced by the SST patch in the west cancels the easterly induced by the SST in the east. Each patch of warm SST enhances evaporation and subsequent deep convection. In other words, the winds tend to cancel each other between the two SST patches. This wind cancellation reduces the intensity of positive feedback and provides an effective negative feedback and acts as a damping force.

In experiment 2, a steep thermocline enhances such damping mechanism. To detect it, we take a snapshot of the model state at the 90th day. Figure 8a shows the anomalous SST and wind fields. Two separate warm SST patches are observed in the eastern ocean along the equator. These two warm SST patches are associated with the first- and second-mode Kelvin waves after the original incident Kelvin wave propagated through the thermocline front. This process is clearly shown in the contours of the anomalous thermocline depth (Fig. 8b). The amplitude of the second Kelvin mode is about 20 meters and the amplitude of the first Kelvin mode is smaller. This agrees with the results of Gill and King (1985) who showed that modal decomposition could be so effective that the second-mode Kelvin wave might have greater amplitude than the first-mode Kelvin wave in the eastern ocean. However, the amplitude of the anomalous SST associated with the first Kelvin mode is greater than that of the second Kelvin mode. The anomalous winds are weaker between these two warm SST patches (Fig. 8a). If there were no such damping mechanism, the anomalous wind would have its maximum amplitude near the center of the SST anomaly. The weakening of the winds over the warm SST associated with the first baroclinic mode reduces the intensity of the air–sea coupling. The SST field in Fig. 8a is similar to the condition shown in Fig. 7.

The condition shown in Fig. 8 is taken from an early stage of the model evolution. In fact, the modal decomposition results from the initial disturbance propagating through the thermocline slope. Does this damping mechanism play a role in the later development of the model oscillation? The modal decomposition and wave reflection are not easy to detect in the
longitude–time plot, such as Fig. 5a of the first experiment. In order to examine this mechanism, we plot the model conditions at the 670th day of the first experiment and at the 760th day of the second experiment when the unstable waves have just propagated through the thermocline slope. The eastern Pacific Ocean, in the first experiment, is dominated by the upwelling process. The thermocline is shallow (Fig. 9b) and the temperature is lower (Fig. 9a). There is a small-amplitude Rossby wave located between $x = 130^\circ$W and $x = 90^\circ$W. The easterly anomalous winds are dominant in the eastern/central Pacific. The easterly winds further intensify the cold event in the eastern ocean. The condition in the second experiment is different. Figure 10b shows the anomalous thermocline depth observed in the second experiment, which has a steeper thermocline profile. Like the first experiment, the upwelling process is dominant in the eastern ocean. However, the amplitudes of the reflected Rossby waves are greater (Fig. 10b). The anomalous SST field is plotted in Fig. 10b. The negative SST is divided into two patches. The easterly winds are absent between these two patches (Fig. 10a), hence, the model continues to decay in this experiment.

It should be pointed out that the decomposition and reflection process seems to dominate in the initial stage.
of coupling and influence the fate of the coupled mode in the further development. This aspect might have some important implications to the initializations of a climate prediction model.

c. The role of western boundary reflection in this model

The work of Neelin (1991) showed the existence of slowly evolving coupled oscillations that are fairly insensitive to the changes of wave speeds. Neelin’s result appears to be very different than those of Battisti (1988), Suarez and Schopf (1988), and Schopf and Suarez (1988), who suggested that the lag time due to propagation of westward Rossby waves and reflected Kelvin waves are of primary importance in determining the oscillation period. We conduct one sensitivity test by removing the solid western boundary and replacing it with an open boundary. Our result (Fig. 11) shows that the oscillatory period remains unchanged although the amplitudes are strongly affected and this result seems to agree with Neelin’s conclusion that several different mechanisms can all contribute to the growth rate of ENSO. It is obvious that the oscillation in this model is closer to the fast-wave limit (the equatorial wave speed is sufficiently fast to bring the ocean into dynamic adjustment rapidly compared to changes due to coupling and temperature advection) which together with the delayed action oscillator represent two different extremes of the same SST mode (Neelin 1991).
It must be pointed out that the results of this sensitivity test are not necessarily contradictory to the concept of the delayed action oscillator proposed by Schopf and Suarez (1988) and Battisti (1988). As pointed out by Neelin (1991), even relatively small changes in the parameter space might change the flow regime. Several assumptions, modifications, and simplifications used in this model may have caused the shifting of the flow regime from the delayed mode. Possible causes will be discussed.

First, a relatively small uniform upwelling $W_0$ is used in this model. In the equatorial Pacific, strong upwelling is confined in the central Pacific and relatively weak upwelling occurs elsewhere. To see the effect of mean upwelling, we linearize the SST equation (3), that is,

$$T_t = -u_w \nabla T_0 - w \frac{T_0 - T_m}{H_m} - \alpha^* T + \sigma(h) h,$$

where

$$\alpha^* = \left( \alpha + \frac{W_0}{H_m} \right), \quad \sigma = \frac{W_0}{h} \frac{\delta T_c}{\delta h} \bigg|_{h=0},$$

and $h$ is the anomalous thermocline depth.

The mean upwelling provides both positive and negative impacts on the coupled instability. A large $W_0$ intensifies the coupling between dynamics and thermodynamics through the coefficient $\sigma$ and enhances the air–sea interaction. But upwelling $W_0$ also increases the effective damping coefficient $\alpha^*$ dramatically. For
example, a relatively small upwelling of 1 m s\(^{-1}\) may introduce an effective damping several times greater than the Rayleigh damping. The positive feedback between the ocean and the atmosphere is essential to the instability growth. The damping mechanism tends to bring the system back to the state of equilibrium. The oscillation results from the competition between these two opposite mechanisms. The use of a uniform upwelling might help to shift the flow regime and cause the difference between the delayed mode and the mode in this model.

Another possible source may be the formulation of subsurface temperature \(T_s\). The formula used by Zebiak and Cane tends to give a weaker coefficient \(\sigma(H)\) especially in the western Pacific where the thermocline is deep. For example, when \(H = 150\) m and \(W_0 = 1\) m s\(^{-1}\), the linearized coefficient \(\sigma(H)\), estimated by the Seager et al. formula, is more than three times larger than that given by Zebiak and Cane's formula.

Besides all possible causes mentioned above, the different treatment of the entrainment term due to the anomalous upwelling [see Eq. (4)] and the exclusion of mean current advection may also help to shift the flow regime.

4. Summary

A tropical atmosphere–ocean coupled model is used to investigate some important effects of a zonally slop-
modes are very sensitive to the steepness of the tilted thermocline in the central Pacific Ocean. The wave reflection and modal decomposition processes play an important role in the development of coupled instability. The insensitivity of the western boundary reflection seems to agree with the results of Neelin (1991) who analytically examined a simplified version of Zebiak and Cane's model and numerically tested a hybrid CGCM and found the western boundary reflection is not of primary importance in the fast-wave limit.

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